Radar Interferometry Detection of Hinge Line Migration on Rutford Ice Stream and Carlson Inlet, Antarctica

Eric Rignot

Jet Propulsion laboratory

Mail Stop 300-243

4800 Oak Grove Drive

Pasadena, CA 91109-8099, U.S.A.

Ph. 818 354-1640; E-mail: eric@adelie.jpl.nasa.gov

Abstract. Satellite synthetic-aperture radar (SAR) interferometry is employed to map the hinge line, or limit of tidal flexing, of Rutford Ice Stream and Carlson Inlet, Antarctica, and detect its migration between 1992 and **1996.** The hinge line is mapped using a model fit from an elastic beam theory. The rms noise of the model fit is -4-10-111111. The hinge line is located with a statistical noise of 30-50 m. Using this method, we find no hinge line migration on Rutford Ice Stream between 1992 and 1996. The measurement noise clue to tide, model fitting, and image registration is 50 m. The possibility that Rutford Ice Stream may be in stable conditions at present is confirmed by the good agreement beteen a revised estimatation of its hinge line ice volume flux and its mass accumulation in the interior. On Carlson Inlet, the hinge line retreats 360±50 m in 4 years along a seaward, southern extension of the hinge line, but is stationary elsewhere. Hence, only a small portion of Carlson Inlet exhibits signs of slow thinning at a rate of 18±3 cm/yr.

1. Introduction.

The region between the grounded part of an ice sheet and the part where the ice begins to float (grounding zone) is of crucial importance when discussing the potential unstability of . the West Antarctic Ice Sheet (WA IS) (Thomas and Bentley, 1978; Hughes, 1977; Weertman, 1974). The grounding line, which is where the glacier detaches from its bed and becomes afloat, may advance seaward as a result of an increase in glacier thickness, an increase in sea level, or an isostatic uprise of the sea-bed. Where the bedrock slopes down toward the center of the ice sheet, as is the case of most of WA IS, the ice sheet may be prone to an irreversible collapse if the grounding line starts to retreat (Thomas, 1979; Hughes, 1977).

Observational data are sparse at the grounding line. Tiltmeters have been used to detect the limit of tidal flexing (Stephenson, 1984), but the method is time-consuming, affected by model assumptions (Smith, 1991), and limited to point measurements. Kinematic Global Positioning System (GPS) is a more effective and accurate technique for locating the limit of tidal flexing (Vaughan, 1995), but spatial sampling is also limited. Satellite visible imagery and radar altimetry offer a larger scale view of the possible location of the grounding line, but at a much coarser spatial resolution. Similarly, radio echo sounding can be used to locate grounding zones, but changes in radio-echo associated with the presence of sea-water undernieth the glacier are not always obvious to interpret (Drewry and others, 1980). As a result of these limitations, most grounding lines are known with considerable uncertainty in the Antarctic.

Recent advances in satellite synthetic-aperture radar (SAR) interferometry suggest that this technique may be able to map the limit of tidal flexing, or glacier hinge line (Holdsworth, 1969), with a precision better than that achieved with any other technique, simultaneously over the entire glacier width, with a uniform sampling scheme (Rignot, 1996). Because the technique is repeatable, it may help detect hinge line migration and thereby provide an early warning of the location and magnitude of glacier changes over the wide areas covered by the satellite.

Here, we employ the radar interferometry technique to map the hinge line of Rutford Ice Stream and Carlson Inlet, Antarctica, using data collected in 1992 and 1996 by ERS-1 and 2 (Fig. 1). These two glaciers drain the Ellsworth Land portion of WAIS into the Weddell Sea, but their flow is restrained by the presence of the Filchner-Ronne Ice Shelves, the world's largest ice shelf system.

The ERS radar data are analyzed here to locate the glacier hinge lines, detect its migration between 1992 and 1996, compare the hinge line ice volume flux with mass accumulation in the interior, and conclude on the state of balance and stability of the glaciers.

2. Study Area

Rutford Ice Stream (RIS) is a fast-flowing outlet glacier which drains 40,500 km^2 of WAIS into the Filchner-Ronne Ice Shelves (Crabtree and Doake, 1982). Two major ground surveys of RIS were conducted in 1978-80 and 1984-1986 (Doake and others, 1987). The surveys showed that the ice stream grounding line has a complex morphology, with several thick ice tongues running along the flow separated by thinner narrow troughs (Stephenson and Doake, 1982). A prominent bedrock knoll is present at the center of the glacier, at the grounding line (Stephenson and Doake, 1982; Vaughan, 1995).

The first radar interferogram of RIS was obtained by Goldstein and others (1993). The deformation signal recorded over the 6-day time interval which separated successive data acquisitions of the ERS satellite was analyzed in terms of its contributions from creep flow, tidal motion, and surface topography. The grounding line of RIS was subsequently mapped with a precision of 500 in. Comparison of this result with Stephenson (1984)'s map suggested a 2 km retreat of the grounding line between 1980 and 1992. The authors however judged the comparison inconclusive because of large uncertainties in registration of the ERS data and in Stephenson's method (Smith, 1991).

Carlson Inlet (CI), north of RIS, only flows at 7 m/yr, compared to 400 m/yr for RIS (Frolich and Doake, 1988). The large contrast in ice velocity was attributed to differences in basal conditions by Frolich and Doake (1989).

3. Method

The premise of using a single radar interferogram to detect the limit of tidal flexing of a glacier was presented by Goldstein and others (1 993). The approach is limited in precision to several hundred meters because the location of the limit of tidal flexing is contaminated by glacial creep flow. '1'o separate glacial creep flow from tidal motion, multiple interferograms are necessary (Hartl and others, 1994). With multiple interferograms, the limit of tidal flexing may then be mapped with a precision of a fcw tens of meters (Rignot, 1996; Rignot and others, 1997).

The hinge line is located upstream of the grounding line (Smith, 1991). Downstream of the grounding line is the line of first hydrostatic equilibrium of the ice, where ice thickness may be inferred from surface elevation. On RIS, the hinge line is separated from the line of first hydrostatic equilibrium by about 2 km, ≈ one glacier thickness (Vaughan, 1995).

4. Interferogram generation

Fourteen ERS scenes acquired in 1992 and 1996 ('l'able 1) *were* utilized to generate 4 interferograms of RIS and CI. The interferograms were subsequently projected onto a polar stereographic (PS) grid with a 50-in sample spacing (Fig. 1). The two 1996 interferograms were combined to estimate both the topography and the line-of-sight velocity of grounded ice using a standard approach (e.g., Zebker and others, 1994). On floating ice, the radar signal is dominated by tidal motion and cannot be used to infer topography and ice velocity.

To help in the process, we utilized a digital elevation model (DEM) of Antarctica assembled from ERS radar altimetry data and distributed on a PS grid at a 5-km spacing (Bamber and Bindschadler, 1997). The DEM was converted from ellipsoid height into orthometric height using the OSU91 geoid model. The DEM provides a highly smoothed version of the local topography, mostly reliable on relatively flat terrain, but nevertheless of sufficient quality to remove - to first order - the effect of surface topography on the interferograms in the proximity of the grounding line. DEM grid points located on the grounded part of the ice streams were used as ground reference to refine (in the least square sense) our initial estimates of the interferometric baseline obtained from the precision ERS orbit data.

Tide-only interferograms were formed by calculating the difference between two interferograms which had been corrected for surface topography using the Antarctic DEM and which spanned the same time interval between data acquisitions (so the creep deformation signal was the same in both interferograms). We assumed that the glacier creep flow was steady and continuous over the time period of observation. If that assumption were violated, we would see deformation fringes on grounded ice which are not clue to topographic errors, baseline uncertainties, or phase noise. We typically do not see such unexplained signal.

One title interferogram was created from the 1992 data and another from the 1996 data. The 1992 and 1996 tide interferograms were co-registered automatically by cross-correlating the radar image intensities on the 1'S grid. Using presumably-stagnant feature points of stable radar characteristics along the ice stream margins, we estimated the precision of registration to be 25 m (half a pixel).

The uncertainty in absolute registration of the imagery is much larger and typically about

200-.500 m. 'l'he error in absolute location of the imagery however does not effect the estimation of glacier changes since the multi-date data are co-registered independently of that information.

5. **Results.**

5.1 Tidal Displacements

Tidal flexure zones stand out in the tide interferograms (Fig. 2 and 3). They correspond to transition regions of high fringe rate which indicate that the glacier surface is vertically displaced over relatively short (few kms) horizontal distances in order to bring the ice into hydrostatic equilibrium. The hinge line is the inward limit of tidal flexing (Holdsworth, 1969). A few km downstream from the hinge line, the fringe rate decreases, which indicates that the glacier surface undergoes less deformation as the ice reaches more stable hydrostatic equilibrium conditions. The flexure zone extends smoothly along the ice margin, confirming that the glacier margins are fully coupled with the grounded ice, as noted earlier by Vaughan (1994).

The shape of the RIS hinge line is consistent with, but probably more accurate than, that inferred by Stephenson (1984) and commented by Smith (1991). The grounded zone of CI exhibits a comparatively sinuous geometry. Both grounding zones differ from straight lines or elevation countour lines. The complexity of the grounding zone is expected from the known rough character of the bedrock topography in this area (Smith and Doake, 1995).

About 60 km downstream from the hinge-line, an isolated zone of tidal flexure is revealed in the middle of Cl, which indicates local grounding of the glacier (Fig.3c-d). The radar brightness of the glacier at that location is high and textured (Fig. 1), which suggests the presence of surface crevasses and an obstruction to ice flow. The flexure zone of the presumed bedrock high occupies nearly half of the glacier width, and must nave a considerable influence on the glacier flow along that channel, perhaps contributing to explain the rather low velocity of CI.

5.2 Comparison with predicted tides

Tidal amplitudes calculated at the bedrock knoll in the center of RIS from four main tidal constituents (Vaughan ,1995) are shown in 'l'able 2 (Vaughan, pers. comm., 1997). The scene to scene variations in tidal amplitude for the 1992 data is three times larger than that calculated for the 1996 data. Indeed, the 1996 scenes are separated by 24 hours, which is close to the repeat cycle of diurnal tides and twice that of semi-diurnal tides; whereas the 1992 data are separated by 6 days, therefore with amplified differences in diurnal and semi-diurnal tides.

The change in tidal amplitude, δz, between different epochs may also be obtained from the tide interferograrns by multiplying the phase values, $\delta\phi$, recorded across the zone of tidal flexure by $\lambda/(4\pi \cos \theta)$, where λ is the radar wavelength (0.0566 m), and $\theta = 23$ degrees is the angle of the radar illumination from the vertical. A single interferogram measures the difference in tidal deformation between two epochs. A tide interferogram measures how that difference varies in between two interferograms, meaning that it measures a quadruple difference in tide. 'l'he interferometric results may therefore be directly compared to quadruple differences in predicted tide.

From the 1992 tide interferogram, we measure a 3.25 m vertical uplift versus 2.97 m predicted. Similarly, the 1996 tide interferogram indicates a -0.96 m vertical uplift versus -0.7 m predicted. The interferograms are therefore in good agreement with the first-order tide predictions.

5.3 Hinge line detection

For each glacier, we selected 4 profiles that cross the zone of tidal flexure (Fig.3). To locate the point of hinging, we utilized a model fit of the profiles based on an elastic beam theory (Holdsworth, 1969). The predicted flexure of an elastic beam, $w(x)$, is written as

$$w(x) = \frac{(w_{max} - w_{min})}{(1 + exp(-\pi))}[1 - exp(-\beta(x - x_0)][\cos(\beta(x - x_0)) + \sin(\beta(x - x_0))], x > x_o \quad (1)$$

where β is the flexural rigidity of the ice (m^{-1}), x is the abscissa along the profile (m), and $x = x_0$ at the hinge line. For each tidal profile, we estimated 4 parameters in the least-square sense: the flexural rigidity of the ice β, the maximum and minimum height of the profile w_{max} and w_{min}, and the position of the hinge line x_o. A measure of the goodness of fit was provided by the rms difference between the model and the interferometric data. The hinge line migration was then measured as the shift in position of x_0 along the profile between 1992 and 1996.

The results, shown in Table 3, indicate a rms noise of the model fit of only a few mm (also see Fig. 5 and 6). This low noise level is remarkable given the model simplicity. Model fitting was however performed only on the segment of the tidal profile closest to the hinge line (meaning extending from 1-2 km upstream of the hinge line, to 4-5 km below the hinge line). Model fitting of longer profiles would obviously not perform as well since the pattern of tidal flexure exhibited by the glacier is not one-dimensional (Fig.2 and 3).

5.4 Comparison with GPS data

A profile P_o was extracted on RIS at the location of a kinematic GPS profile collected in 1993 (A-A' in Vaughan, 1995). The GPS and ERS profiles compare well, although the GPS data exhibits considerably more noise (Fig. 4). A model fit of the GPS data yields a rms noise of 20 mm, one order of magnitude larger than that achieved with interferometry (3 mm). The comparison suggests that radar interferometry measures tidal displacements better than GPS, and thereby offers a level of precision in mapping of the glacier hinge line which is totally unprecedented.

To obtain a good overlap between the GPS and ERS curves, we had to shift our interferegram 500 m to the West on the PS grid. The GPS data are geolocated with 5 m precision (Vaughan, pers. comm., 1997). We would not expect the hinge line to migrate 500 m in 3 years based on the results discussed in the next section. The shift is therefore due to an error in absolute geolocation of the ERS imagery. To eliminate that error, control points of known geolocation must be found in the radar imagery.

5.5 Hinge line migration

To determine the precision with which wc measure hinge line migration, the model fit comparison was extended on both sides of the selected profiles P_i's over an area 500 m wide (or 1/3 of the glacier thickness), which means that 5 parallel profiles were examined on each side of the main profile. The values of x_o were calculated for each one of the 11 profiles and the results were averaged to calculate a mean offset and a standard deviation.

The standard deviation of the offsets represents the statistical noise of detection of hinge line migration. Its value averages 30 m ('l'able 3).

I'refiles P_1 and P_3 on RIS exhibit no hinge line migration (Table 3). P_1 runs across the bedrock knoll in the center of the glacier which is believed to pin down the location of the grounding line (Fig.3a). The glacier slope at that location is 0.03 (Vaughan, pers. comm., 1997) compared to O.OO2 over the rest of the glacier. Hence, a hinge line migration is least likely to be observed over that portion of the glacier. Profile P_3 suggests that the northern margin of the glacier is stable. Profile P_2 on the southern flank indicates a 69 m retreat. The only significant shift is a 130-111 advance along P_4. Overall, these numbers are relatively small and suggest that the hinge line of RIS did not migrate between 1992 and 1996. Converted to ice thickness changes using a mean slope of 0.002, the measured hinge line migration suggests that the thickness of RIS did not change by more than 5 cm/yr.

On CI, the hinge line advanced 139 m along P_1, remained the same along P_2 and P_4, but ret rested 360 m along P_3 (Fig.6). P_3 runs along a seaward, southern extension of the grounding zone. The measured shift is large, and may correspond to an 18-cm/yr thinning of the glacier assuming a glacier slope of 0.002.

5.6 Influence of tide

The stability of the RIS hinge line is somewhat of a surprise because changes in tide should affect to some degree the position of the hinge line. At high tide, the glacier may lift up from its bed and displace the hinge line inward. Conversely at low tide, the hinge line may move seaward. Depending on the geometry of the glacier bed at the hinge line and other factors (for instance the ice temperature at the glacier bed), either effects or none may influence the position of the hinge line.

The largest positive tide experienced by RIS during the ERS data acquisition was -0.03 m in 1992 and -0.386 m in 1996 (q'able 2). The largest negative tide was -1.826 In in 1992 and -1.471 m in 1996. The differential title (which is the only one that matters for detecting hinge line migration) is 0.355 m in both cases. Assuming a glacier slope of 0.002, this sea-level differential should move the hinge line inward by 177 m. We do not observe a systematic shift of that order of magnitude in our data.

The absence of a 177-m hinge line migration on both glaciers suggest that the effect of tide is much less than that predicted from the glacier slope. Both glaciers must be anchored rather solidly onto rough bedrock, with limited migration of the hinging point during tidal cycles.

Most likely, the above statement is simplistic and tide does effect the location of the hinge line. Since about 3/4 of the profiles exhibit a hinge line migration of ± 50 m, wc suggest that the effect of tide is to induce a ± 50-m uncertainty in position of the hinge line. With this level of uncertainty, the hinge line of both glaciers would appear stable along all profiles except one. The hinge line retreat of CI along P_3, which is the only one exceeding that predicted from tide, must correspond to a real thinning trend of the glacier which should deserve further investigation.

Mass balance of Rutford Ice Stream

Radar interferometry only measures the ice velocity in the line of sight of the radar. In the absence of data collected from different orbits, the line-of-sight velocity may be combined with estimates of flow direction to produce two-dimensional velocity estimates. Here, we estimated flow direction from flow features present in the radar imagery along the ice stream margins and linearly interpolated in between. The resulting velocity profile of RIS about 2 km seaward from the hinge line is shown in Fig 7.

The ice flux of RIS was computed at that location assuming an ice with a density of 890 kg/m^3 in hydrostatic equilibrium over sea-water with a density of 102IS kg/m^3 (Crabtree and Doake, 1982). The error in ice velocity normal to the profile is 5 percent due to uncertainties in flow direction. Ice thickness is known to within 5-10 percent from the precision of the DEM (l-2 m). The ice flux should therefore be accurate to 10 percent. We measure 13.1 ± 1 km3/yr for RIS.

Crabtree and Doake (1982) estimated the ice flux of RIS to be 20.4 ± 2 km3/yr (or 18.5 Gt/yr water equivalent) based on one velocity estimate (400 m/yr) and one estimate of ice thickness (1.7 km). Our estimate is probably more accurate since it is based on an entire velocity and thickness profile.

Using an accumulation area of 40,500±4000 km^2 from Crabtree and Doake (1982), the balance accumulation of RIS is 0.32±0.03 m/yr ice equivalent. The accumulation data of Giovinetto and Bentley (1985), recently revised and re-gridded on a 50-km grid (Giovinet to, pers. comm. 1997) reads 0.33 m/yr ice equivalent in the accumulation area of RIS. This value is within 10 percent of our calculated balance accumulation, which itself is known with 10 percent uncertainty. The comparison therefore suggests that RIS is presently close to a state of mass balance.

Conclusions

The example discussed in this study illustrates the usefulness of radar interferometry for providing essential information on glacier grounding line stability and mass balance in the Antarctic. The technique provides a complete mapping of the glacier hinge line with unprecedented precision (a few tens of meters). Radar interferometry may also be used to detect hinge line migration.

On Rutford Ice Stream, we find that the hinge line is remarkably stable. The result seems consistent with the state of mass balance of the glacier indicated by a revised estimate of its hinge line ice discharge and re-gridded estimates of its mass accumulation. On the neighboring, slow-moving Carlson Inlet, the hinge line shows signs of glacier thinning along a southern, seaward extension of the grounding zone.

To confirm these results and improve on their precision, a solution is to examine additional interferograms. With additional data, we should be able to constrain the (not-well known) effect of title and isolate the long term hinge line migration more effectively.

Acknowledgements I would like to thank R. Dietrich and D. Vaughan for their thoughtful reviews of the manuscript and for their numerous suggestions to improve its contents. I also thank D. Vaughan for kindly providing tide predictions and GPS profiles of Rutford Ice Stream to help constitute an invaluable set of ground truth for this study. I thank J. Bamber for providing us with a DEM of Antarctica without which this study would have been considerably more difficult. Finally, I thank the European Space Agency for making the ERS data available to us and for operating such a superb mission. This work was performed at the Jet Propulsion Laboratory, California Institute of Technology, under a contract with the National Aeronautics and Space Administration.

References.

Bamber, J. and R.A. Bindschadler. 1997. An improved elevation data set for climate and ice sheet modelling: validation with satellite imagery, *Ann.Glaciol.* 25, in press.

Crabtree, R.D. and C.S.M. Doake. 1982. Pine island glacier and its drainage basin: results from radio echo-sounding, *Ann. Glaciol.* 3, 65-70.

Doake, C. S. M., Frolich, R. M., Mantripp, D.R., Smith, A. M., ancl D.G. Vaughan. 1987. Glaciological studies on Rutford Ice Stream, Antarctica, *J.Geophys.Res.*, 92(119), 8951-8960.

Drewry, D. J., D.T. Meldrum and E. Jankowski. 1980. Radio echo and magnetic sounding of the Antarctic ice sheet, 1978-79, *Polar Rec.* 20(124), 43-51.

Frolich, R.M. and C.S.M. Doake. 1988. Relative importance of lateral and vertical shear on Rutford Ice Stream, Antarctica, *Ann. Glaciol.* 11, 19-22.

Frolich, R. M., Vaughan, D.G. and C.S.M. Doake. 1989. Flow of Rutford Ice Stream and comparison with Carlson Inlet, Antarctica, Ann. *Glacio.* 12, 51-56.

Giovinetto, M. Il. and C.R. Bentley. 1985. Surface balance in ice drainage systems of Antarctica, *Antarct. J. U.S.* 20(4), 6-13.

Goldstein, R. M., Engelhardt, H., Kamb, B., and R.M. Frolich. 1993. Satellite radar interferometry for monitoring ice sheet motion: application to an Antarctic ice stream, Science 262(5139), 1525-1530.

Hartl, P., K.H. Thiel, X. Wu, C. Doake and J. Sievers. 1994. Application of SAR interferometry with ERS-1 in the Antarctic. *Earth observation Quarterly* 43, ESA Pub., 1--4.

Holdsworth, G. 1969. Flexure of a floating ice tongue. *J.Glaciol.*, 8(54), 385-397.

Hughes, T.J. 1977. West Antarctic ice streams, Rev. *Geophys. Space Phys.* 15(1), 1-46.

Jacobs, S. S., H.H. Helmer, C.S.M. Doake, A. Jenkins and R.M. Frolich. 1992. Melting of ice shelves and the mass balance of Antarctica. *J. Glaciol.* 38(130), 375-387.

Rignot, E. 1996. Tidal flexure, ice velocities and ablation rates of Petermann Gletscher, Greenland, *J.Glaciol.*, 42(142), 476-485.

Rignot, E., S. P. Gogineni, W.B. Krabill and S. Ekohlm. 1997. Ice discharge from north and northeast Greenland as observed from satellite radar interferometry, Science, 276, 93'1-937.

Robin, G. de Q., Glaciology 111, seismic shooting and related investigations, in *Norwegian-*

British-Swedish Antarctic Expedition, 1949-52, Scientific Results, Vol. I. Norsk Polarinstitut, Oslo, 1958.

Smith, A.M. and C.S.M. Doake, Seabed depths at the mouth of Rutford Ice Stream, Antarctics, *Ann. Glaciol.*, 20, 353-356, 1995.

Smith, A.M. 1991. 'l'he use of tiltmeters to study the dynamics of Antarctic ice-shelf grounding lines. *J. Glaciol.*, 37(125), 51-58.

Stephenson, S.N. and C.S.M. Doake. 1982. Dynamic behavior of Rutford Ice Stream. *Ann. Glaciol.* 3, 295-299.

Stephenson, S.N. 1984. Glacier flexure and the position of grounding line: measurements by tiltmeter on Rutford Ice Stream, Antarctica, *Ann. Glaciol.*, **5**, 165 169.

Thornas, R.H. and C.R. Bentley. 1978. A model for Holocene retreat of the west antarctic ice sheet, *Quarternary Research* 10(2), 15-170.

Thomas, R. Il. 1979. The dynamics of marine ice sheets, *J. Glaciol.* 24(90). 167 177.

Vaughan, D.G. 1994. Investigating tidal flexure on an ice shelf using kinematic GPS. *Ann. Glaciol. 20, 372-376.*

Weertman, J. 1974. Stability of the junction of an ice sheet and an ice shelf. *J. Glaciol.* 13(67), 3-11.

Zebker, H.A., P.A. Rosen, R.M. Goldstein, C. Werner and A. Gabriel. 1994. On the derivation of coseismic displacement fields using differential radar interferometry: the Landers Earthquake. *J. Geophys. Res.*, 99 (1110), 19,617-19,634.

Table 1. ERS SAR scenes used in this study. $B_\perp$ is the interferometric baseline in the direction perpendicular to the line of sight of the radar. Frame 5337 corresponds to Rutford Ice Stream. Frame 5319 corresponds to Carlson Inlet. ERS is a sun synchronous satellite orbiting at about 800 km altitude, illuminating the surface at 23 degrees away from the vertical, with a 100-km swath, at a spatial resolution of 4 m in the along track direction (azimuth) and 20 m in the cross-track direction (range).

Orbit Pair	Frames	Date(yy-mm-dd/dd)	$B_\perp$ (m)
23600(ERS1)/3927(ERS2)	5319-5337	96-01-02/03	210
22598(ERS1)/2925(ERS2)	5319-5337	95-11-10/03	74
2943/3029	5319-5337	92-02-07/13	60/-138
3029/3115	5319-5337	92-02-13/20	-138

Table 2. Tide predictions for Rutford Ice Stream, Antarctica at the location of the bedrock knoll (Vaughan, pers. comm., 1997).

Orbit	Date(yy-doy), Time (hr-mn)	Tide (Pred.) (m) (m)
23600(E1)	96-019 11:02	-0.386
3927 (E2)	96-020 11:02	-0.813
22598(E1)	95-315 11:02	-1.471
2925 (E2)	95-316 11:02	-1.198
2943 (E1)	92-038 11:04	-1.826
3029 (E1)	92-044 11:04	-0.030
3115 (E1)	92-050 11:04	-1.204

Table 3. Hinge line migration (m) determined from 4 profiles by model fitting of the interferometrically-determined tidal displacements. The measurement precision (±) is determined by varying the position of the profile by ±250 m along the north axis. The rms difference (mm) between the model fit and the data is indicated in parenthesis. The profiles are located as shown in Fig.2. The surface slope of the glacier on the grounded part of the profile (proxy indicator of ice thickness slope) calculated from the Antarctic DEM is 0.002 for all profiles except P_1 on RIS which is 0.03 (Vaughan, pers. comm., 1997).

Profile	Rutford Ice Stream	Carlson Inlet
P_1	-13±9 (7)	139±32 (13)
P_2	-69±26 (5)	-56±33 (11)
P_3	31±42 (6)	-360±9 (11)
P_4	130±11 (9)	-48±38 (-)

Figure Caption

Figure 1. Radar amplitude image (120 x 220 km) in a polar stereographic projection of (a) Rutford Ice Stream and (b) Carlson Inlet acquired by ERS-1 on Jan. 1, 1996. ©ESA1996.

Figure 2. Tidal motion of (a) Rutford Ice Stream and (b) Carlson Inlet in 1996 after removal of the surface topography using a DEM of Antarctica (Bamber and Bindschadler, 1997). Residual fringes on grounded ice are caused by artefacts in the DEM. Each full cycle of grey tone variation represents an 88.4-mm increment in vertical displacement of the glacier surface induced by tidal motion. White squares show the locations of Figure 3a-d.

Figure 3. Tidal motion of (a) Rutford Ice Stream in 1992; (b) Rutford Ice Stream in 1996; (c) Carlson Inlet in 1992; and (d) Carlson Inlet in 1996 on a polar stereographic grid with a 50-m sample spacing. The profiles P_i's refer to Table 3 and Fig. 4 and 5. Each full cycle of grey tone variation represents a 62-mm vertical displacement of the glacier surface. To facilitate the comparison between the 1992 and 1996 interferograms, the 1996 tidal displacements were multiplied by 3.505 to approximate the amplitude of the 1992 tidal displacements. The tidal displacements recorded on grounded ice near the hinge line average zero with a rms noise of 3 mm.

Figure 4. Comparison of a GPS profile acquired in 1993 (Vaughan, 1995) and tidal displacements measured in 1996 by ERS (profile P_o in Fig. 3a). The rms of the model fit (continuous curve) is 20111111 for the GPS and :1111111 for ERS.

Figure 5. (a) Model fitting (continuous line) of the radar interferometry data (dotted line) acquired along profile P_1 of Rutford Ice Stream (Fig. 3a) in 1996 (black dots) and 1992 (grey dots); (b) difference between the 1996 and 1992 profiles (squares) and model fits (dotted curve). No data are available in the middle of the profile due to high phase noise in the 1992 interferogram. The rms noise of the model fit is 6 mm for the 1996 data and 4 mm for the 1992 data. The inferred flexural rigidity of the ice, β, is 0.24 km^{-1} in both cases. The hinge line migration is -2 m.

Figure 6. Model fitting (continuous line) of the radar interferometry data (dotted line) acquired along profile P_3 of Carlson Inlet (Fig. 3c). The rms noise of the model fit is 8 mm (1996) and 11 mm (1992). The inferred flexural rigidity of the ice, β, is 0.2S km^{-1}. The hinge line migration is -345 m.

Figure 7. Ice velocity parallel to the glacier surface along the flow direction (continuous thick, dark line), ice velocity normal to the grounding line profile (here discretized into linear segments) (dotted, dark, thick line) and ice thickness deduced from hydrostatic equilibrium of the ice (continuous, thick, grey line) on Rutford Ice Stream in 1996.

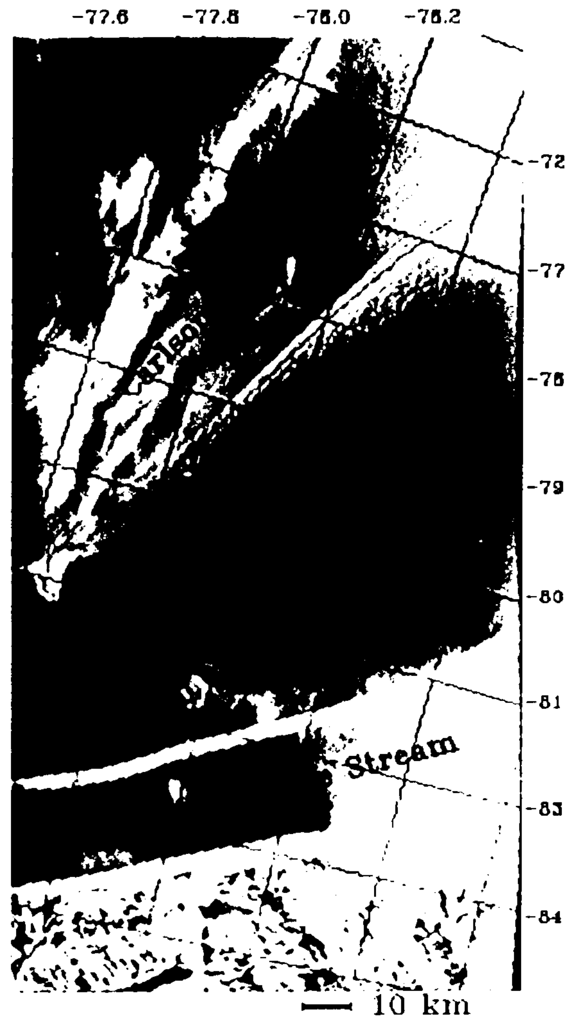


Figure 1 Radar amplitude image (120 x 220 km) in a polar stereographic projection of Rutford Ice Stream and Carlson Inlet acquired by ERS 1 on Jan. 1, 1996. ©ESA 1996.

(a) Rutford Ice Stream

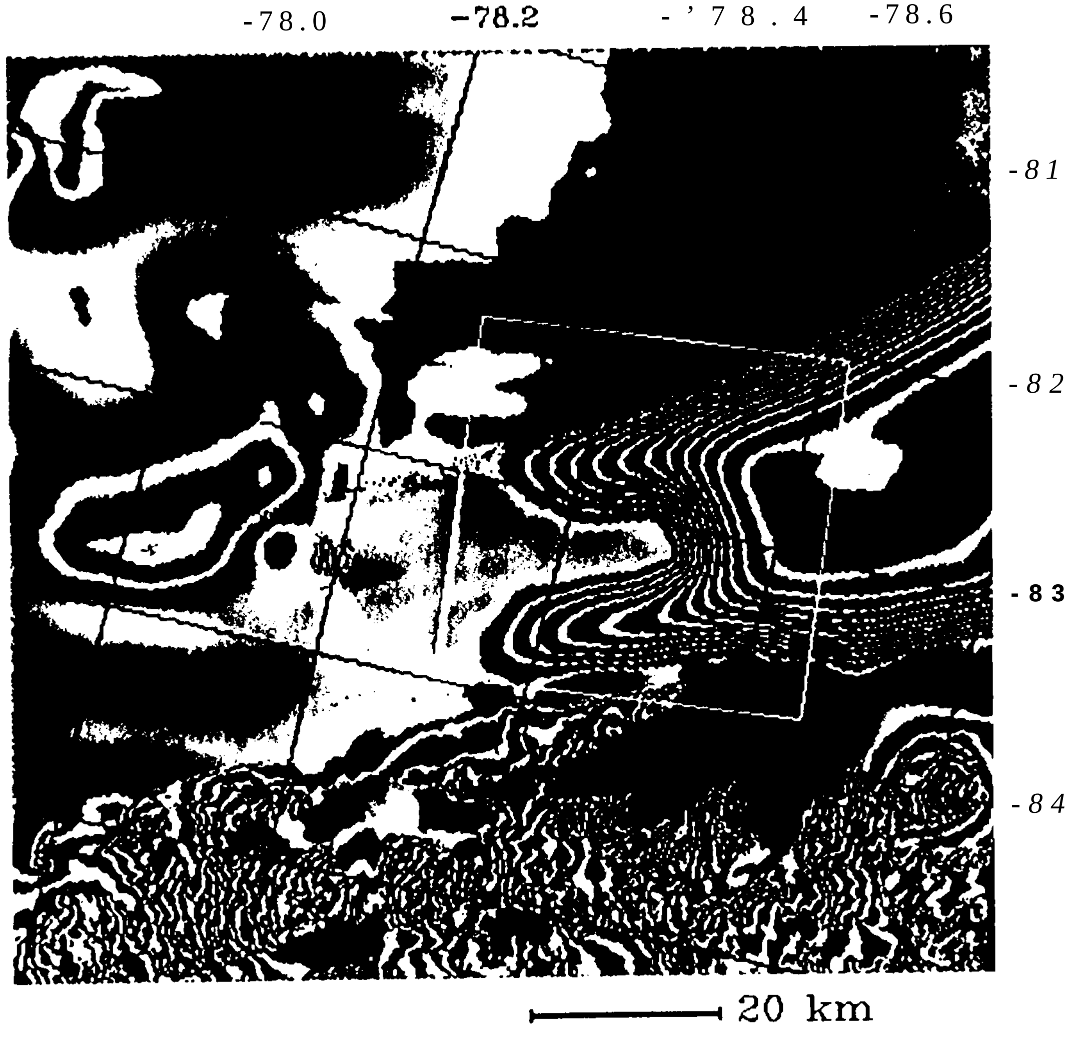
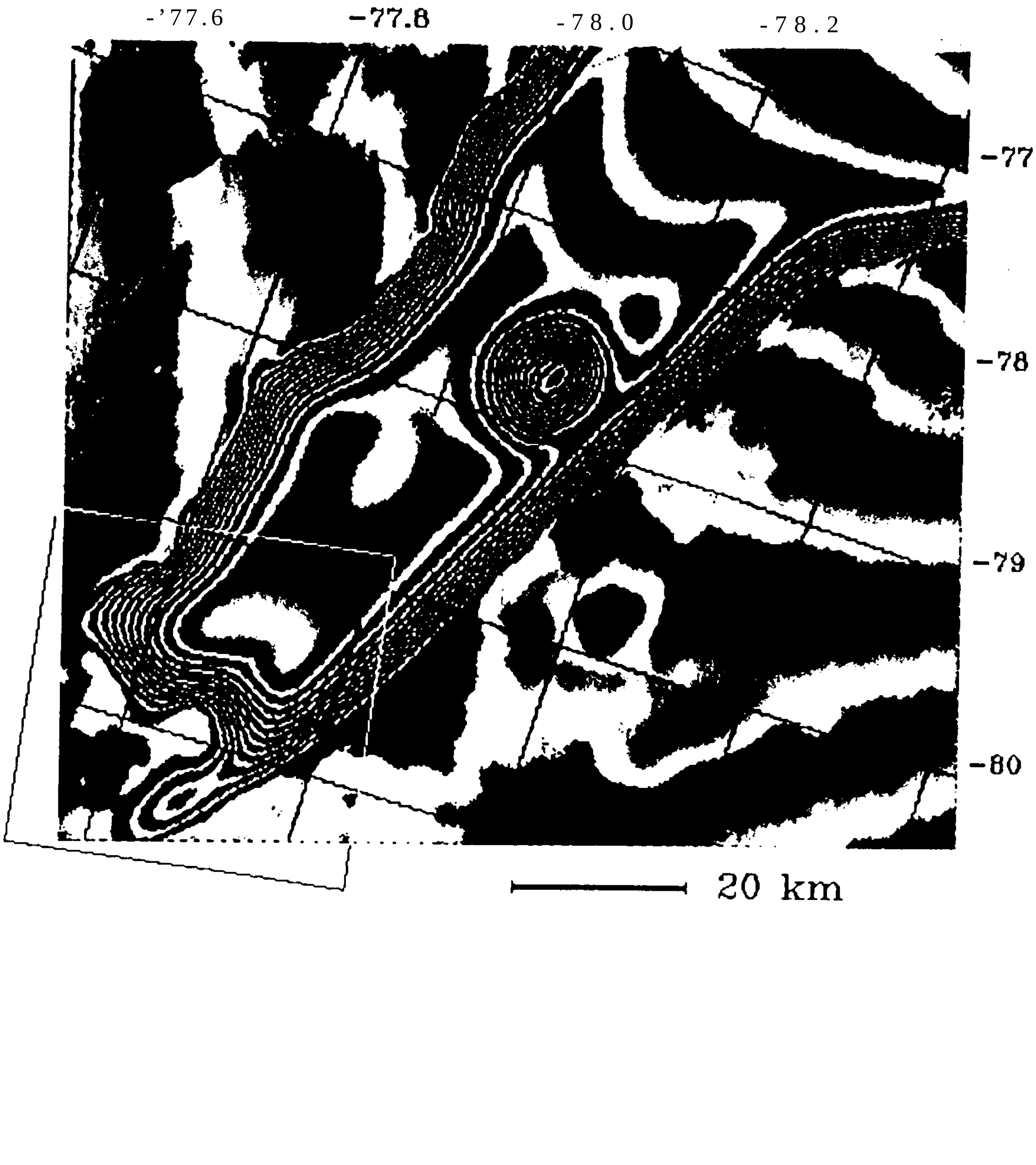
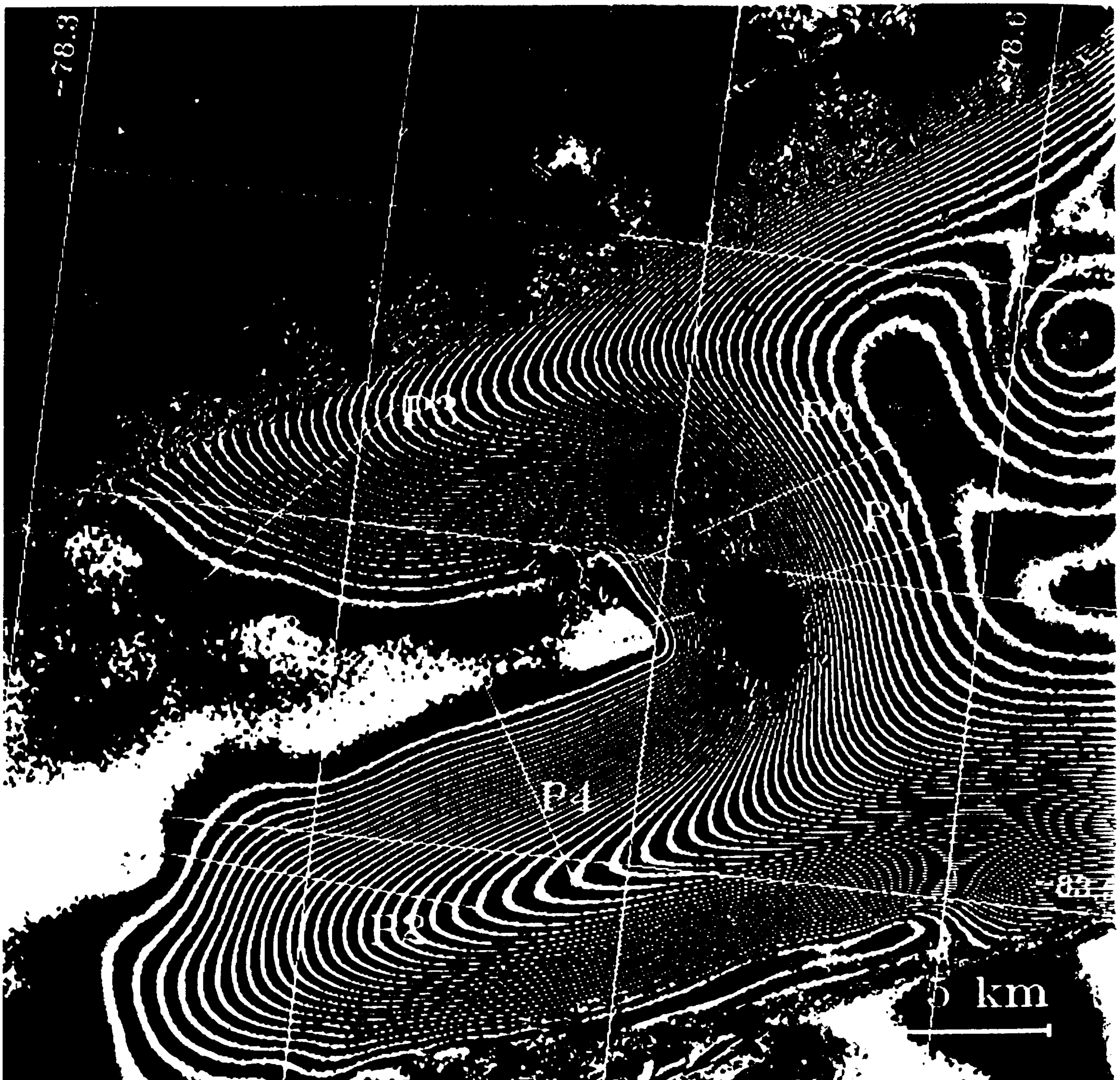


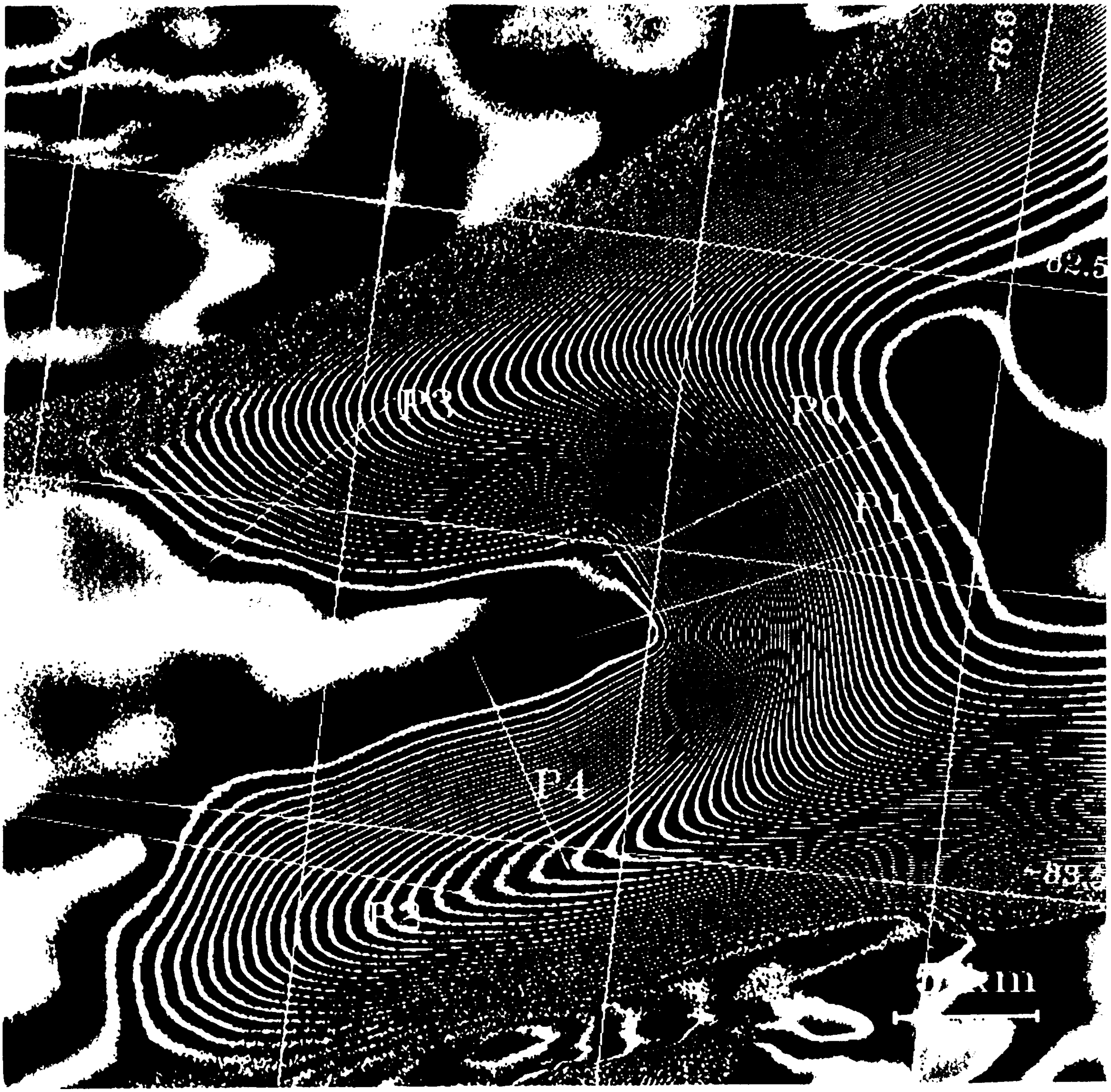
Figure 2. Tidal motion of (a) Rutford Ice Stream and (b) Carlson Inlet in 1996 after removal of the surface topography using a DEM of Antarctica (Bamber and Bindschadler, 1997). Residual fringes on grounded ice are caused by artefacts in the DEM. Each full cycle of grey tone variation represents an 88.4-mm increment in vertical displacement of the glacier surface induced by tidal motion. White squares show the locations of Figure 3a-d.

(b) Carlson Inlet

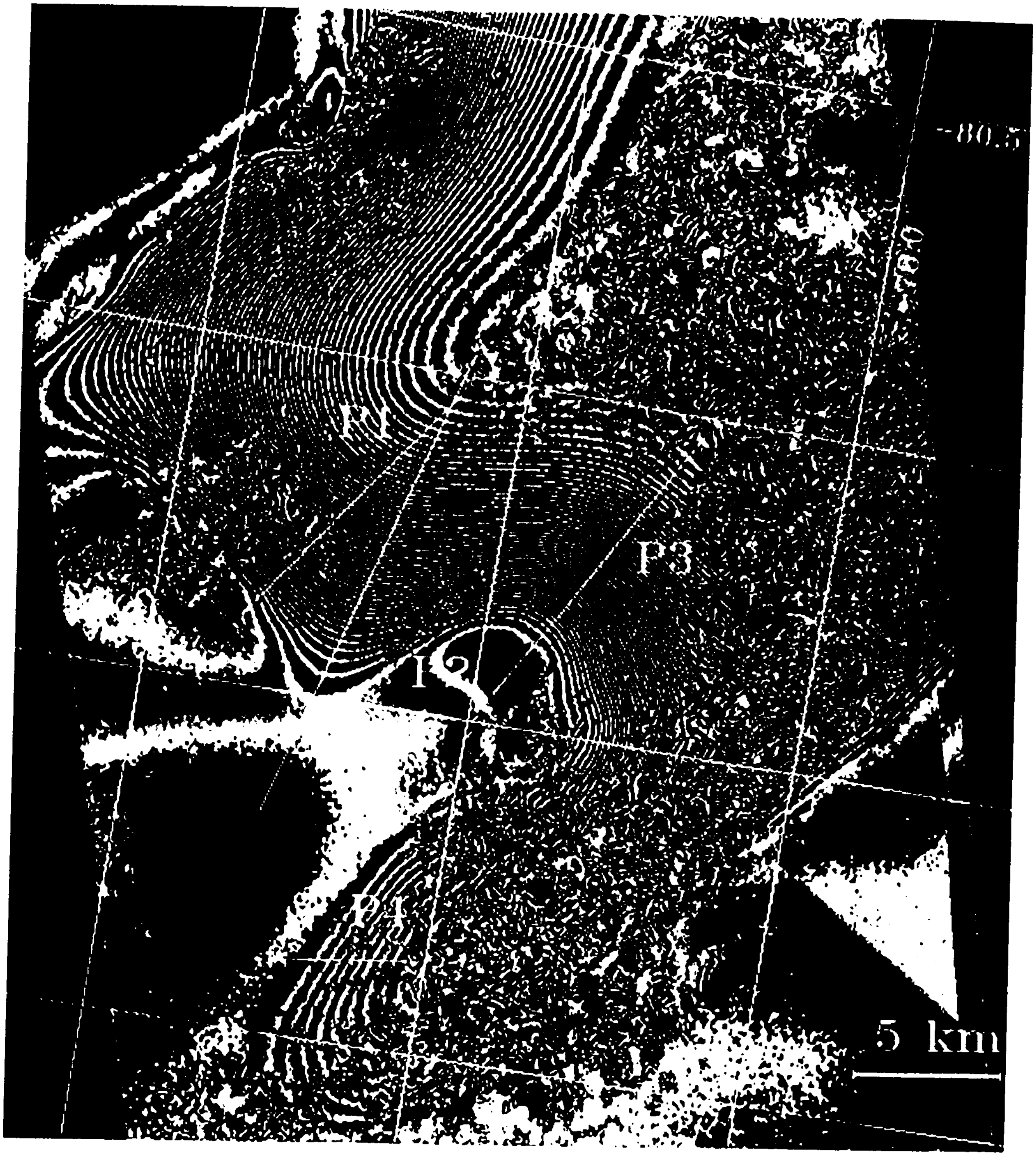




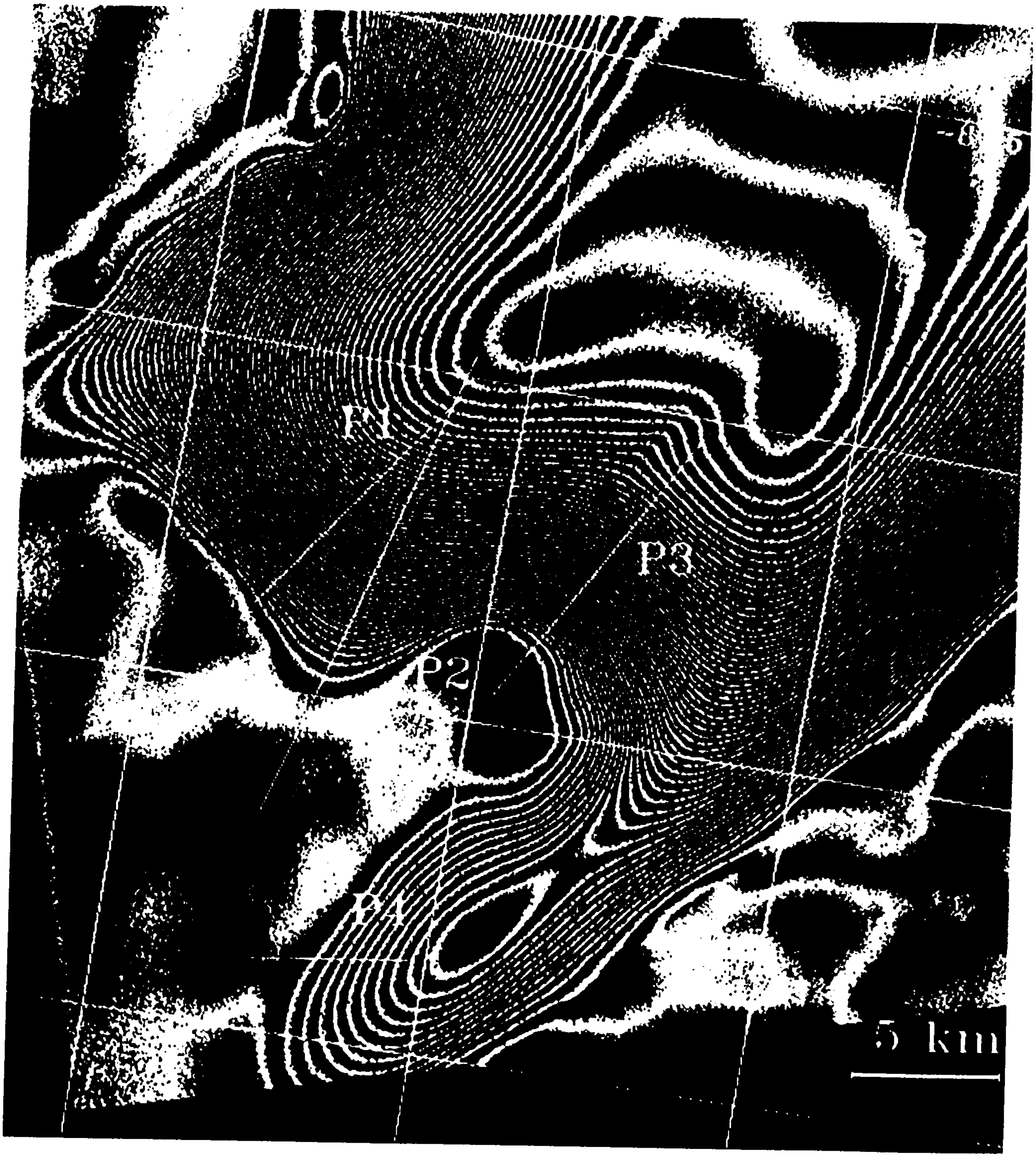
-78.3
-78.6
P3
P6
P1
P4
P2
-83
5 km


P3
P0
P1
62.5
P4

(c) Carlson Inlet in 1992



(d) Carlson Inlet in 1996



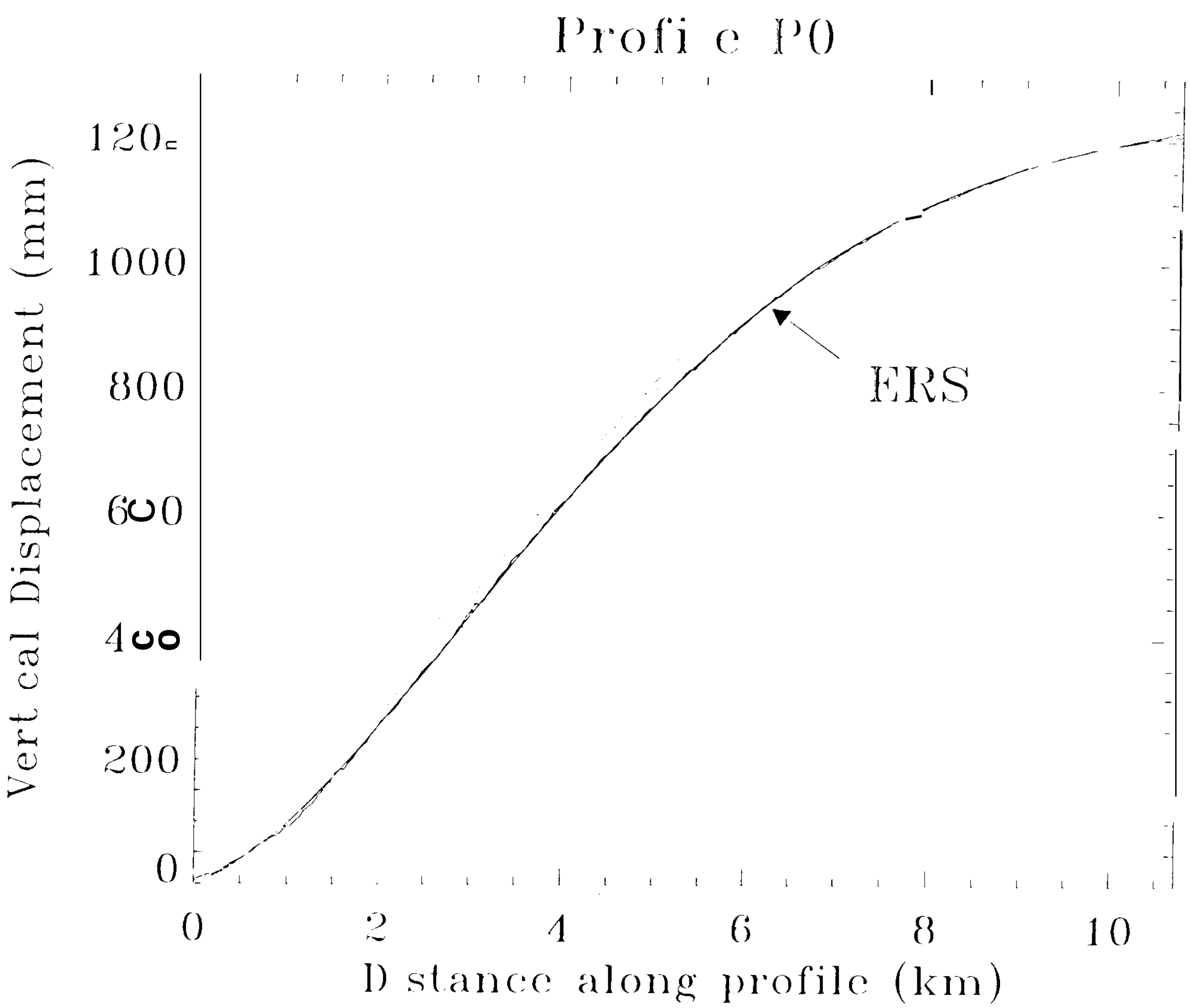


Figure 4. Comparison of a GPS profile acquired in 1993 (Vaughan, 1995) and tidal displacements measured in 1996 by ERS (Profile P_0 in Figure 3). Model fits for the two curves have a rms difference of 20 mm for GPS and 3 mm for ERS

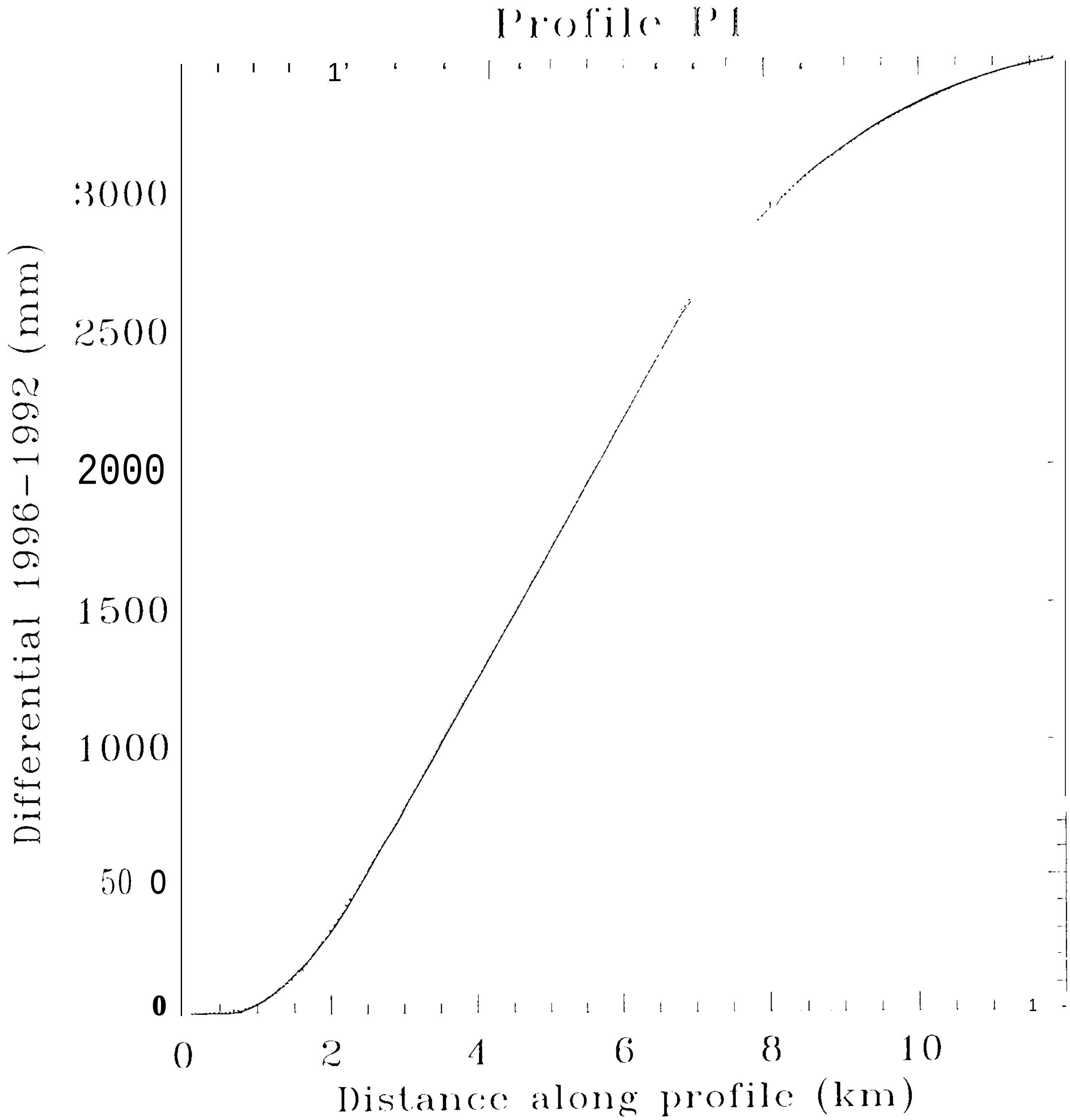


Figure 5. (a) Model fitting (continuous line) of the radar interferometry data (dotted line) acquired along profile P_1 of Rutford Ice Stream (Figure 3a) in 1996 (black dots) and 1992 (grey dots).

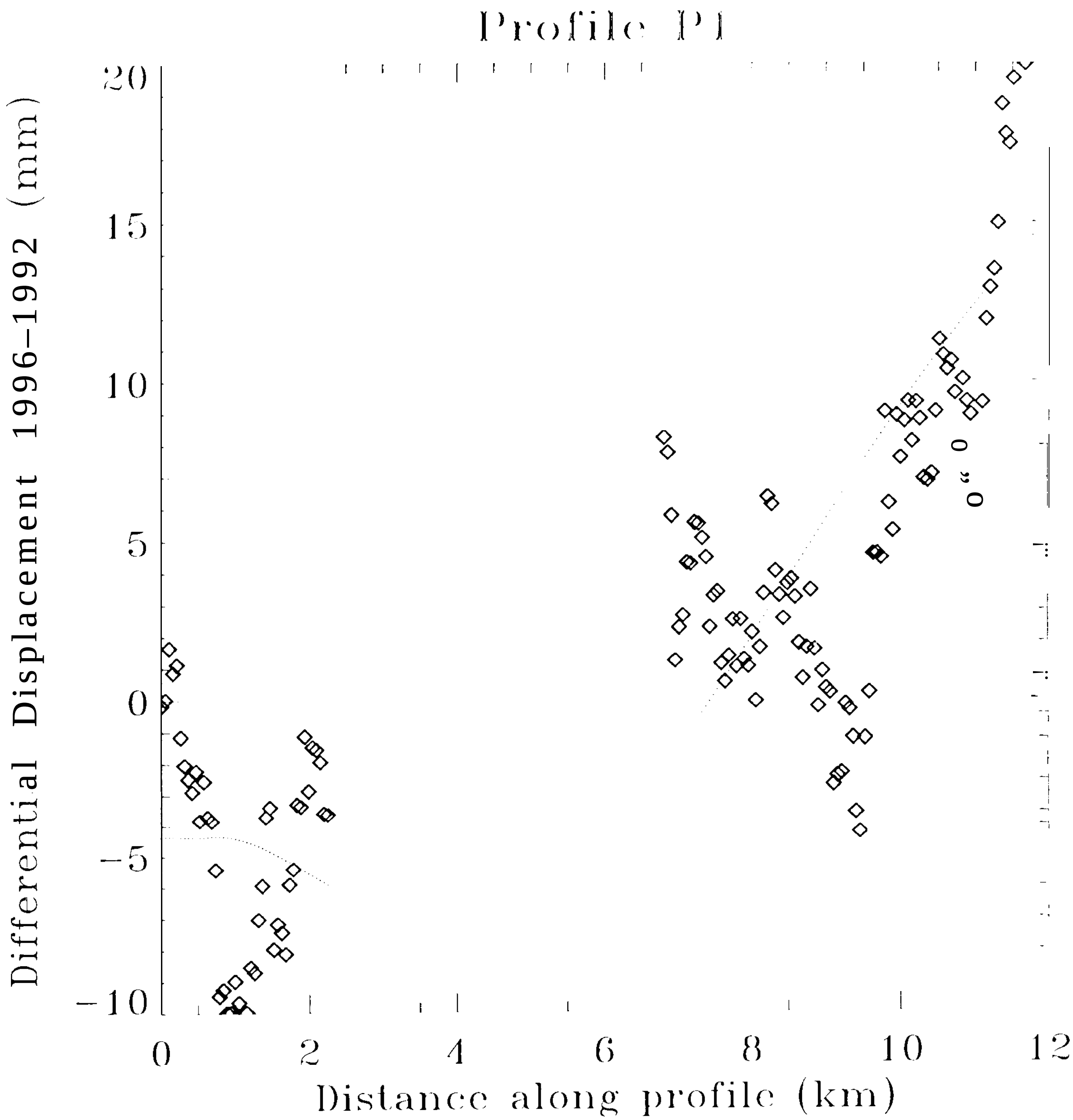


Figure 5. (b) difference between the 1996 and 1992 profiles (squares) and model fits (dotted curve). The rms differences between the model and the interferometry data are 6 mm (1996) and 4 mm (1992). The inferred flexural rigidity of the ice, β, is 0.24 km^{-1} in both cases. The hinge line migration is -2 m.

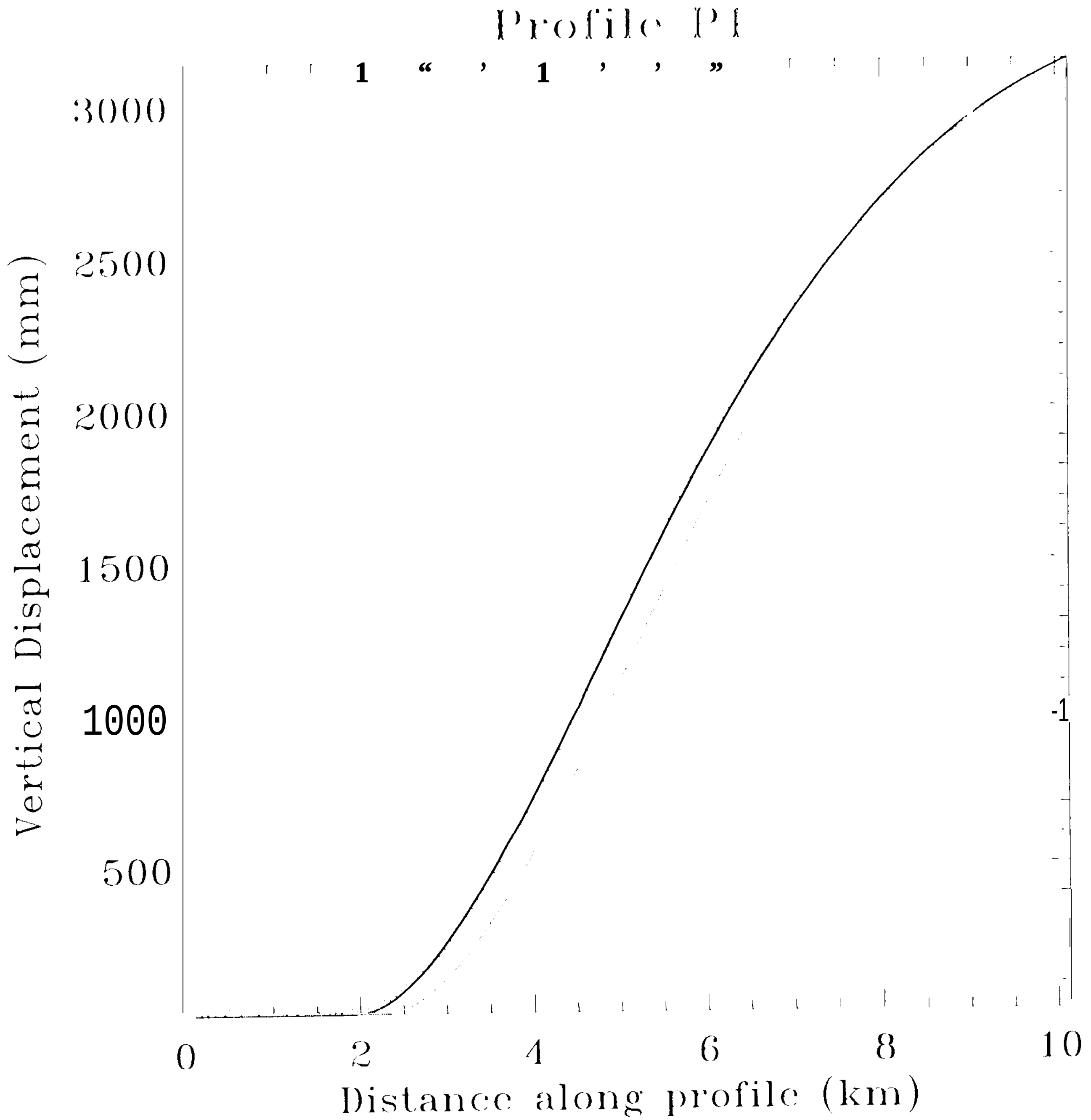


Figure 6. Model fitting (continuous line) of the radar interferometry data (dotted line) acquired along profile P_3 of Carlson Inlet (Figure 3c). The rms differences between the model and the interferometry data are 8 mm (1996) and 14 mm (1992). The inferred flexural rigidity of the ice, β, is 0.28 km^{-1} in both cases. The hinge line migration is -345 m

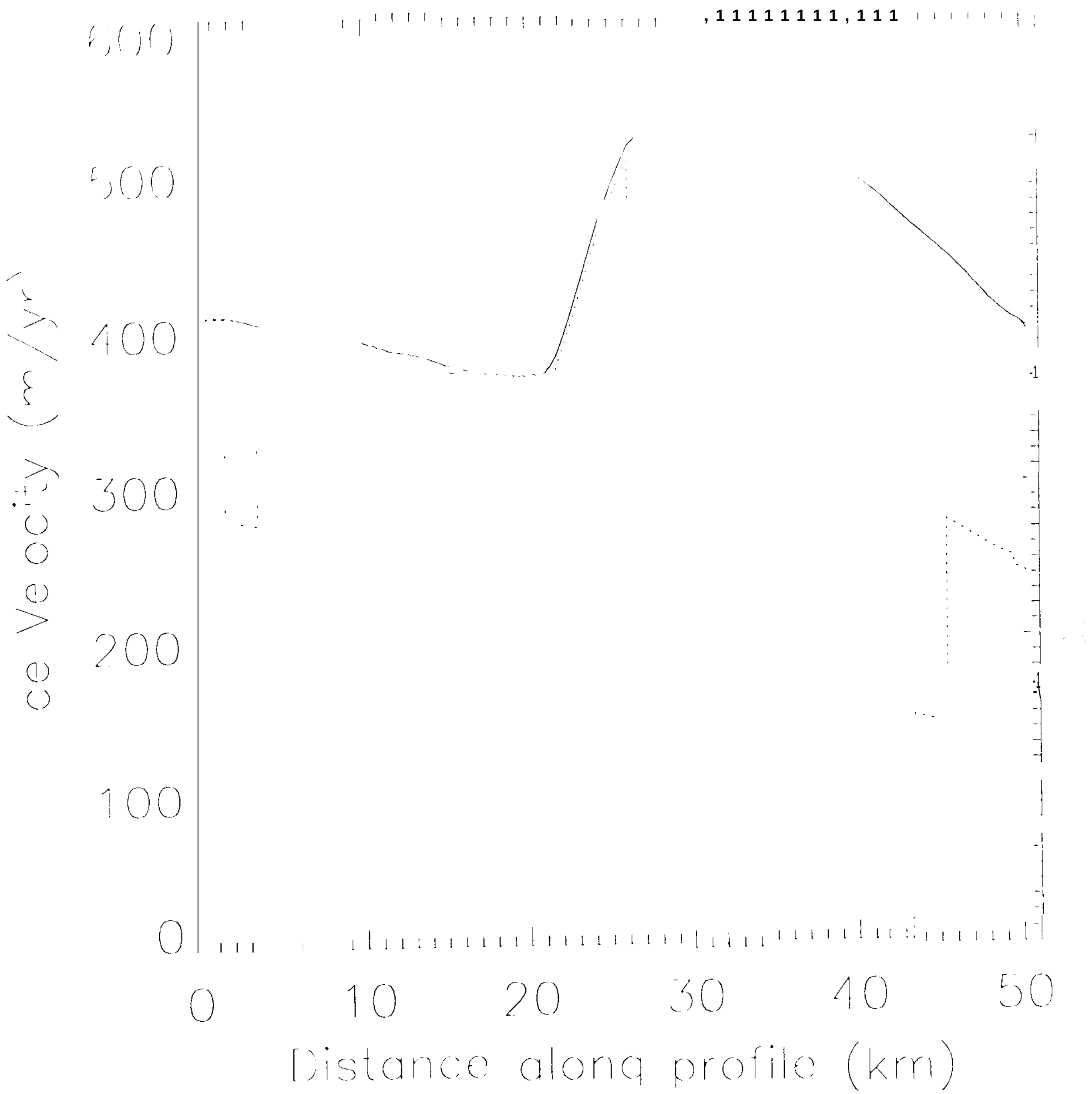


Figure 7. Ice velocity parallel to the glacier surface in the flow direction (continuous thick, dark line), ice velocity normal to the grounding line profile (discretized in linear segments) (dotted, dark, thick line) and ice thickness deduced from hydrostatic equilibrium of the ice at the grounding line (continuous, thick, grey line) on Rutford Ice Stream in 1996.

Dear Editor,

I am thankful for the reviewers' comments and in response to these I have extensively modified the manuscript. Response to detailed comments are given below. The main features of the revision is an improvement of the figures (lat-lon coordinates), the addition of in-situ measurements (tide and GPS), and a more complete evaluation of error sources and precision of the method.

Response to Reviewer 2 (**Reinhard Dietrich**)

Remark 1 (DEM): The DEM of Antarctica used in this study is rectified from ellipsoidal heights into orthometric heights using the OSU91 geoid model. We added a sentence in the text to describe this correction.

Remark 2: Yes, tide predictions could be useful and I incorporated such information in the revised manuscript.

Response to Reviewer 1 (**David** Vaughan)

Note: David gave very useful comments, and also provided the author with an extremely valuable data set which included GPS tidal profiles collected in 1993 on Rutford Ice Stream and tide predictions for Rutford Ice Stream at the time of passage of the ERS satellite.

* Response to main comments:

I have derived a *more explicit error budget*. I added lat-lon on figure 1 and others as well as scales. Ice thickness scale was added on last figure.

The english is not perfect, true. I tried to improve it as best as I can. For the final version, I will ask a collegue to proof-read it and make necessary corrections. Unfortunately, my english wilt never be as impeccable as that of an englishman. I apologize for the inconvenience to the reader.

The revised manuscript was submitted in double space. My apologies for not doing so in the first place.

* Response to suggestions:

I modified the title. I put capitals on Rutford Ice Stream and Carlson Inlet.

True that "Seaward grounding line migration would result from isostatic uplift".

I took out "entrain".

I noted the surface slope values quoted by David, especially that of the bedrock knoll which was not correctly represented in our DEM of Antarctica.

Anandakrishnan 1997 described how seismic waves generated at the grounding line could propagate upstream of the grounding line. 1 cannot see now this would mean that the glacier velocity also fluctuates with tidal cycles at the grounding line.

The DEM of Antarctica is indeed only a coarse representation of the topography over the ice streams, but it is sufficient to remove the main surface slope of the glacier from the interferometric data. Of course, the DEM will not remove the small scale surface topography, , but slnall-scale topography is not of major significance to the determination of the limit of tidal flexing.

The velocity of the ice streams is derived independent of the DEM because in that easel derived the topography from the radar data (only on grounded ice ! On floating ice the method cannot work because of tidal contamination of the radar signal).

Figure 3a has been removed. 1 did not discuss the details of how two interferograms are used to measure topography and ice velocity of grounded ice because the text is already long, the main object ive is to discuss tides, and the description of the method for evaluating topography and velocity on grounded ice has been discussed elsewhere by the present author as wc]] as collegu es, so there is nothing really new on that matt er. On Rut ford Ice Stream, BAS collected an extensive GPS data set which would be useful for validation purposes but beyond the scope of this st udy.

In the revised manuscript 1 compared one of Vaughan's GPS profiles with ERS. As the ot]ıer profiles gave similar results they were not included in the discus sion.

1 noted in tile revised manuscript that Vaughan (1994) had already hinted that the ice margin was fully coupled with the ice shelf.

Since the grounding line is unlikely to migrate at the bedrock knoll, I compared the position of the hinge line on 4 profile's on each glacier, including along the flanks of the knoll on Rutford Ice Stream.

1 hope David will find the paper more suitable for publication with all these additions. His help was in any case greatly appreciated.